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PREDICTING SEDIMENT TRANSPORT BY INTERRILL OVERLAND FLOW ON ROUGH SURFACES

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ABSTRACT

Modelling soil erosion requires an equation for predicting the sediment transport capacity by interrill overland flow on rough surfaces. The conventional practice of partitioning total shear stress into grain and form shear stress and predicting transport capacity using grain shear stress lacks rigour and is prone to underestimation. This study therefore explores the possibility that inasmuch as surface roughness affects flow hydraulic variables which, in turn, determine transport capacity, there may be one or more hydraulic variables which capture the effect of surface roughness on transport capacity sufficiently well for good predictions of transport capacity to be achieved from data on these variables alone. To investigate this possibility, regression analyses were performed on data from 1506 flume experiments in which discharge, slope, water temperature, rainfall intensity, and roughness size, shape and concentration were varied. The analyses reveal that 89·8 per cent of the variance in transport capacity can be accounted for by excess flow power and flow depth. Including roughness size and concentration in the regression improves that explained variance by only 3·5 per cent. Evidently, flow depth, when used in combination with excess flow power, largely captures the effect of surface roughness on transport capacity. This finding promises to simplify greatly the task of developing a general sediment transport equation for interrill overland flow on rough surfaces.

KEY WORDS: sediment transport; soil erosion; interrill flow; overland flow; hillslopes

INTRODUCTION

The sediment transport capacity of overland flow is the maximum flux of sediment that a flow is capable of transporting. Both the rate of erosion (i.e. soil detachment and removal) and the rate of deposition by flowing water are controlled by the difference between the transport capacity and the influx of sediment from upslope, with erosion occurring where this difference is positive and deposition where it is negative (e.g. Foster and Meyer, 1972; Foster *et al.*, 1995; Smith *et al.*, 1995). Because transport capacity controls both the rate of erosion and the rate of depostion, it is a property of fundamental importance in the quantitative representation of the processes of soil erosion and deposition. Consequently, virtually all physically based soil erosion models developed during the past two decades contain a sediment transport capacity equation (e.g. Foster and Meyer, 1975; Beasley *et al.*, 1980; Woolhiser *et al.*, 1990; Smith *et al.*, 1995; Foster *et al.*, 1995; Morgan *et al.*, 1998). The ability of these and future models to provide accurate predictions of soil erosion is therefore dependent on the accuracy of the sediment transport capacity equation they employ.

Many existing soil erosion models use either a bed load or total load formula originally developed for rivers as their transport capacity equation (e.g. Foster and Meyer, 1975; Woolhiser *et al.*, 1990; Smith *et al.*, 1995). Other soil erosion models utilize simple empirical formulas in which transport capacity is related to a measure of flow intensity (e.g. Gilley *et al.*, 1985; Hartley, 1987; Foster *et al.*, 1995; Morgan

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et al., 1998). All these formulas have been developed and/or calibrated in flumes with plane beds (i.e. beds without large-scale roughness). This is a major weakness because overland flow on hillslope surfaces, whether natural or disturbed by agriculture, is characterized by large-scale roughness (i.e. roughness that disturbs the water surface), and this roughness might be expected to significantly reduce the sediment transport capacity.

In river flow the traditional means of dealing with the effect of bedforms on sediment transport capacity is to partition the total shear stress into grain shear stress and form shear stress, and to predict transport capacity using grain shear stress (e.g. Einstein, 1950; Carson, 1987). Govers and Rauws (1986) and Govers (1988) presented evidence suggesting that a similar approach may work in overland flow. However, Govers (1992) pointed out that widely differing predictions may be obtained in overland flow depending on the method employed to calculate grain shear stress. Using Govers and Rauws' (1986) method of calculating this quantity, Abrahams and Parsons (1994) and Atkinson *et al.* (1998) demonstrated that grain shear stress grossly underpredicts sediment transport capacity in overland flow. The reason appears to be that most turbulent eddies generated by bedforms in deep river flow occur away from the bed (Einstein, 1950), whereas in overland flow eddies produced by large-scale roughness elements occur so close to the bed that they have a profound effect on sediment transport.

In an early soil erosion model, Foster (1982) distinguished between shear stress acting on the soil and that acting on the surface cover (i.e. plants, mulch and microtopography). This distinction is incorporated into the rill erosion module of the WEPP (Water Erosion Prediction Project) model (Foster *et al.*, 1995). Because this approach uses experimental data to estimate the amount by which surface cover slows the flow and, hence, reduces the transport capacity, it is superior to the traditional grain shear stress approach. Nonetheless, it suffers from the problem that it is impossible to evaluate experimentally the effect on flow velocity of every kind and density of surface cover. In addition, there are no comparable experimental data available for interrill flow, which is the focus of this study.

Yet another approach to predicting the transport capacity of overland flow has been suggested by Govers (1992). He proposed using hydraulic variables that implicitly account for the effect of bed roughness. This approach is based on the speculation that inasmuch as surface roughness affects flow hydraulics which, in turn, determines transport capacity, there may be one or more hydraulic variables which capture the effect of surface roughness on transport capacity sufficiently well for good predictions of transport capacity to be achieved from data on these variables alone. Two hydraulic variables that Govers (1992) thought might be suitable were Yang's (1972) unit flow power and Govers' (1990) effective flow power. Govers' idea is incorporated into EUROSEM (the European Soil Erosion Model) (Morgan *et al.*, 1998) which uses effective flow power to predict the transport capacity of interrill flow on rough surfaces. The predictive equation is based on Everaert's (1991) experiments on plane beds, but it has not been tested against data from rough beds. Thus, it is by no means clear that the equation in particular or the approach in general works. The present study was therefore undertaken to investigate this approach using data from both plane and rough beds. If good predictions of transport capacity can be achieved from hydraulic variables alone, the problem of developing a sediment transport equation for interrill flow on rough surfaces becomes more tractable.

SETUP AND METHODS

The flume employed in this study was 5·2m long and 0·4m wide with a smooth aluminium floor and plexiglas walls (Figure 1). It consisted of two parts: a lower part, 3·6m long, which was covered with sand, and an upper steeper part, 1·6m long. For the experiments the lower part of the flume was inclined at slopes β of 2·7°, 5·5° and 10·0°. A well sorted silica testing sand with a median diameter D of 0·74mm (ASTM C-190) and a density ρ_s of 2650kg m^{-3} was supplied by a continuously adjustable sediment feed system to the upper part of the flume and was trapped at the lower end of the flume in two containers. Water entered the upper end of the flume by overflowing from a head tank. This inflow was controlled by a gate valve and measured with a paddlewheel flow meter (OMEGA model FP-5300) on the inlet pipe

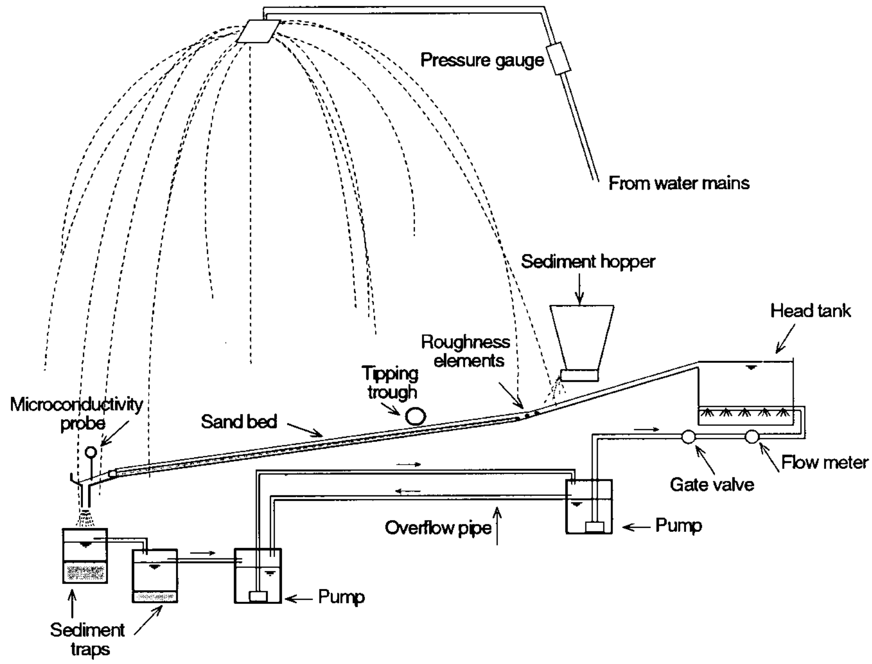

Figure 1. Sketch of the flume and rainfall simulator. When simulated rainfall was used, the water flowing through the flume was not recirculated

to the head tank. The water discharge Q ($m^3 s^{-1}$) from the flume was equal to the inflow, except in those experiments involving simulated rainfall where Q was equal to the sum of the inflow and the rainfall. During each experiment, the sediment feed rate was adjusted to Q so that the bed was experiencing no perceptible scour or deposition. Sediment concentration C_s (kg m^{-3}) was determined by sampling the water–sediment mixture leaving the flume.

Use of this type of flume is based on the assumption that water entering the upper end of the flume will pick up all the sediment it is capable of transporting before it reaches the lower end of the flume. The purpose of the sediment feed system is merely to replace the sand being removed from the bed and so prevent scour from exposing the flume floor. Govers (1990) measured transport capacities in 3m and 6m long flumes of this type (without a sediment feed system) and concluded that they gave comparable results. Everaert (1991) studied transport capacities of interrill overland flow in a flume that was only 0·3m long. Consequently, it is believed that the 3·6m length of the sand-covered section of the present flume was sufficient for transport capacities to be achieved.

The mean flow velocity u (m s^{-1}) was determined by a salt tracing technique described elsewhere (Li and Abrahams, 1997). Knowing Q and u, mean flow depth d (m) was calculated using

$$d = Q/uw \tag{1}$$

and

$$w = W(1 - C) \tag{2}$$

Table I. Summary of experimental data

	1	2	3	4	5	6
Roughness type	Plane	Plane	Cylinders	Stones	Stones	Miniature trees
Number of experiments	75	106	620	237	394	74
Reynolds number	2074–12 546	2084–11 290	2018–18 618	2113–21 434	2006–16 643	2321–18 405
Froude number	1·04–2·73	0·89–2·88	0·29–1·61	0·38–2·10	0·56–1·93	0·16–1·12
Slope (degrees)	2·7, 5·5, 10	2·7, 5·5, 10	2·7, 5·5, 10	5·5	2·7, 5·5, 10	2·7, 5·5, 10
Sediment size (mm)	0·74	0·74	0·74	0·74	0·74	0·74
Roughness concentration			0·04–0·35	0·04–0·37	0·08–0·34	0·10–0·28
Roughness diameter (cm)			0·95. 2·16. 3·17. 8·90	2·80, 4·55, 9·13	2·80, 4·55, 9·13	0·10
Rainfall intensity (mm h^{-1})		53–158			53–152	
Kinematic viscosity ($\times 10^{-6} m^2 s^{-1}$)	1·053–1·138	0·979–1·224	0·788–1·514	0·955–1·037	0·896–1·197	1·096–1·340
Bedload (%)	70·4–98·4	72·1–96·6	46·8–94·8	53·8–84·4	58·1–86·9	37·1–78·9

where w is the flow width (m), W is the flume width (m), and C is the concentration of roughness elements defined as the proportion of the bed covered by these elements. The density of the water ρ was assumed to be 1000 kg m^{-3}, while the kinematic viscosity v ($m^3 s^{-1}$) was determined from water temperature. Fixing ρ while varying v with temperature can be justified on the grounds that ρ varies by less than 0·05 per cent over the measured range of temperature while v varies by a factor of almost 2.

Because the mechanics of sediment transport in laminar overland flow (Li and Abrahams, 1998) are different from those in transitional and turbulent overland flow, the experiments were confined to transitional and turbulent flow and had Reynolds numbers $Re = 4ud/v$ ranging from 2006 to 18 618. Froude numbers $F = u/(gd)^{1/2}$ varied between 0·16 and 2·88, where g is the acceleration of gravity (m s^{-2}). A summary of the experiments is given in Table I. SI units are used in all computations and analyses in this study.

Altogether 1506 experiments were performed in six series. The first and second series were conducted on a plane bed with no roughness elements. During the second series simulated rainfall was applied at target intensities of 54, 108 and 162 mm h^{-1} from one, two and three Spraco-Lechler full cone jet nozzles (Luk *et al.*, 1986) mounted 0·3 m apart and 3·6 m above the centre of the flume. Actual intensities I determined from eight rain gauges departed slightly from the target values. The coefficient of variation for the eight gauges averaged 0·093 over nine 30 min events, indicating a remarkably uniform spatial distribution. The median drop sizes (Laws and Parsons, 1943) associated with the target intensities were 1·6, 2·0 and 2·4 mm, respectively. Drop size increased with rainfall intensity because the spray cones intersected when the second and third nozzles were operating, causing the water drops to collide and form larger drops. Given that the flow velocity from each nozzle exceeded 5 m s^{-1}, most drops reached terminal velocity. From the terminal velocity–drop size relation (Laws, 1941; Gunn and Kinzer, 1949) and the measured drop size distributions, the total kinetic energies of the three intensities were computed to be 0·24, 0·65 and 1·19 J m^{-2} s^{-1}, respectively. These kinetic energies are respectively 48, 68 and 85 per

cent of the energies of natural rainstorms at the same intensities (Kinnell, 1973).

The third series of experiments employed cylinders with four different diameters D_r as roughness elements. The smallest cylinders were made from wooden dowls 0·95cm in diameter, and the remaining cylinders from polyvinyl chloride (PVC) pipe with diameters of 2·16, 3·17 and 8·90cm. The cylinders were arranged by eye into random patterns and were sufficiently tall that they were never inundated by the flow. In a special effort to obtain a wide range of v in this series, water temperatures were varied from 2·5 to 32°C. This was done to ascertain whether viscosity influences sediment transport through its effect on suspended sediment.

In the fourth and fifth series the flume bed was covered with stones to mimic a desert hillsope. Three groups of stones were collected from a local river bed and the lengths of their long a, intermediate b, and short c axes measured. The stones in the three groups were similar in shape (with a median Corey shape factor $c/(ab)^{1/2}$ close to 0·6) and roundness (with a modal roundness of 5 on Powers' (1953) six-point scale) but were different in size. The stones were randomly placed in the flume with their a–b planes parallel to the bed. The median values of $(a + b)/2$ for the three groups were 2·80, 4·55 and 9·13cm. These values were taken to represent D_r. Because the stones were partly buried in the sandy bed, the smallest stones were inundated by most flows, the intermediate-sized stones by only the highest flows, and the largest stones by none of the flows. Simulated rainfall was used in the series 5 but not in the series 4 experiments.

In the series 6 experiments miniature artificial Christmas (i.e. conifer-like) trees were employed as roughness elements. The trees provided a form of roughness quite different from the cylinders and stones and more akin to grass, plant stems and litter. The randomly distributed trees were about 9cm wide and 11cm high and had wire branches and plastic leaves 0·1cm thick. Although the trees impeded the flow, water was still able to pass through them. Thus, the actual roughness concentration C was less than the apparent concentration C_a viewed from above. To estimate C from C_a, flow depths were measured for several high discharges and compared with flow depths calculated from $d_a = Q/uw_a$, where $w_a = W(1-C_a)$. Based on these comparisons, C was set equal to $0{\cdot}4C_a$ in all experiments. The use of a constant correction factor of 0·4 was crude but unavoidable. As this factor almost certainly varied from one experiment to another depending on flow depth (because the trees were roughly conical in shape) and tree selection (because no two trees were the same), the hydraulic data are less reliable for this series than for the other five. Finally, given that the flow passed between adjacent branches and leaves, the selected value for D_r was the branch and leaf thickness of 0·1cm.

Bed sediments may be transported as bedload (i.e. by rolling and saltation) or suspended load (i.e. supported by turbulent eddies). Hu and Hui (1996) developed an equation for predicting the proportions of such sediments in these two modes of transport. However, the equation was derived from plane-bed experiments and may not be an accurate predictor of these proportions in the present experiments (i.e. series 3 to 6) characterized by large-scale roughness. Nevertheless, in the absence of any better methodology, the equation is used here to estimate the bedload percentage in each experiment. The data reported in Table I signify that in the vast majority of the experiments, bedload exceeds suspended load and in some cases accounts for more than 90 per cent of the total load.

APPROPRIATE VARIABLES

Measure of transport capacity

An important issue in a study such as this is whether to represent sediment transport capacity by a measure of sediment concentration or sediment load. There are two reasons for employing sediment concentration in this role. First, concentration is arguably a more fundamental variable than sediment load, insofar as sediment load is the product of concentration and discharge. Second, the use of sediment concentration rather than sediment load avoids the possibility of spurious correlations with hydraulic variables selected to predict transport capacity. However, there is a major disadvantage in using

sediment concentration where one is dealing with overland flow on rough surfaces. As Figure 2 illustrates, where the size, shape and concentration of roughness elements, the energy slope and the sediment size are fixed, sediment concentration C_s varies conservatively with discharge Q. In fact, in most situations C_s remains virtually constant (i.e. within the range of measurement error) as Q increases. Thus C_s correlates poorly with discharge and, indeed, with any hydraulic variable correlated with discharge. Consequently, in this study transport capacity is represented by a measure of sediment load.

There are two measures of sediment load in common usage: the dry sediment transport rate

$$T_d = C_s Q / w \tag{3}$$

which has units kgm^{-1}s^{-1}, and the immersed sediment transport rate

$$T_w = T_d g \rho_s / (\rho_s - \rho) \tag{4}$$

which has units kg s^{-3} or W m^{-2}. Although the former measure is the simpler of the two and has been widely employed in soil erosion studies, the latter is physically more meaningful in that it corrects for buoyancy and has the dimensions and quality of the rate of doing work (Bagnold, 1966). In the analysis of the present data, it makes little practical difference which measure is employed. Therefore, on theoretical grounds alone T_w is used as the measure of sediment transport capacity.

Measures of hydraulic conditions

In overland flow on rough surfaces the basic hydraulic variables controlling sediment transport are S, d, u, ρ, ρ_s and v, where $S = \sin \beta$. Of these, the most important are S, d and u (insofar as ρ and ρ_s are constant and v does not correlate with T_w in the present data set). S, d and u may be employed separately as predictor variables in a sediment transport equation or they may be combined in a variety of ways to form composite predictor variables. Three such composite variables are mean bed shear stress $\tau = g\rho dS$, Yang's (1972) unit flow power uS, and Bagnold's (1966) specific flow power $\omega = \tau u$. Table II shows that regardless of whether S, d and u are employed separately to predict T_w or are combined in τ, uS, or ω, the predictive power of these variables as measured by the coefficient of determination R^2 and the standard error of estimate (*SEE*) is the same. Therefore, from a practical perspective it makes no difference which combination of these variables is selected to represent hydraulic conditions. However, from a theoretical point of view, a case can be made for choosing a combination that includes ω on the grounds that the appropriate flow quantity on which T_w depends is in the nature of a power supply or rate of energy dissipation per unit area of the bed, and ω is the relevant measure of this quantity. What is more, ω has the same units as T_w. Actually, the variable that is used in the present analysis is excess flow power $\omega - \omega_c$, where the subscript c denotes the critical value of the variable at which sediment begins to move. Excess flow power is used because when bed sediments are being transported at capacity, the shear stress at the bed borne by the fluid as opposed to the grains is just equal to the critical shear stress (e.g. Owen, 1964; Bagnold, 1973). Thus, the power available to transport the grains is $\omega - \omega_c$ rather than ω.

ANALYSIS

Stepwise multiple regression was utilized to investigate whether sediment transport capacity of interrill overland flow on rough surfaces can be well predicted without the use of variables characterizing surface roughness. The data were obtained from the six series of experiments described above. The dependent

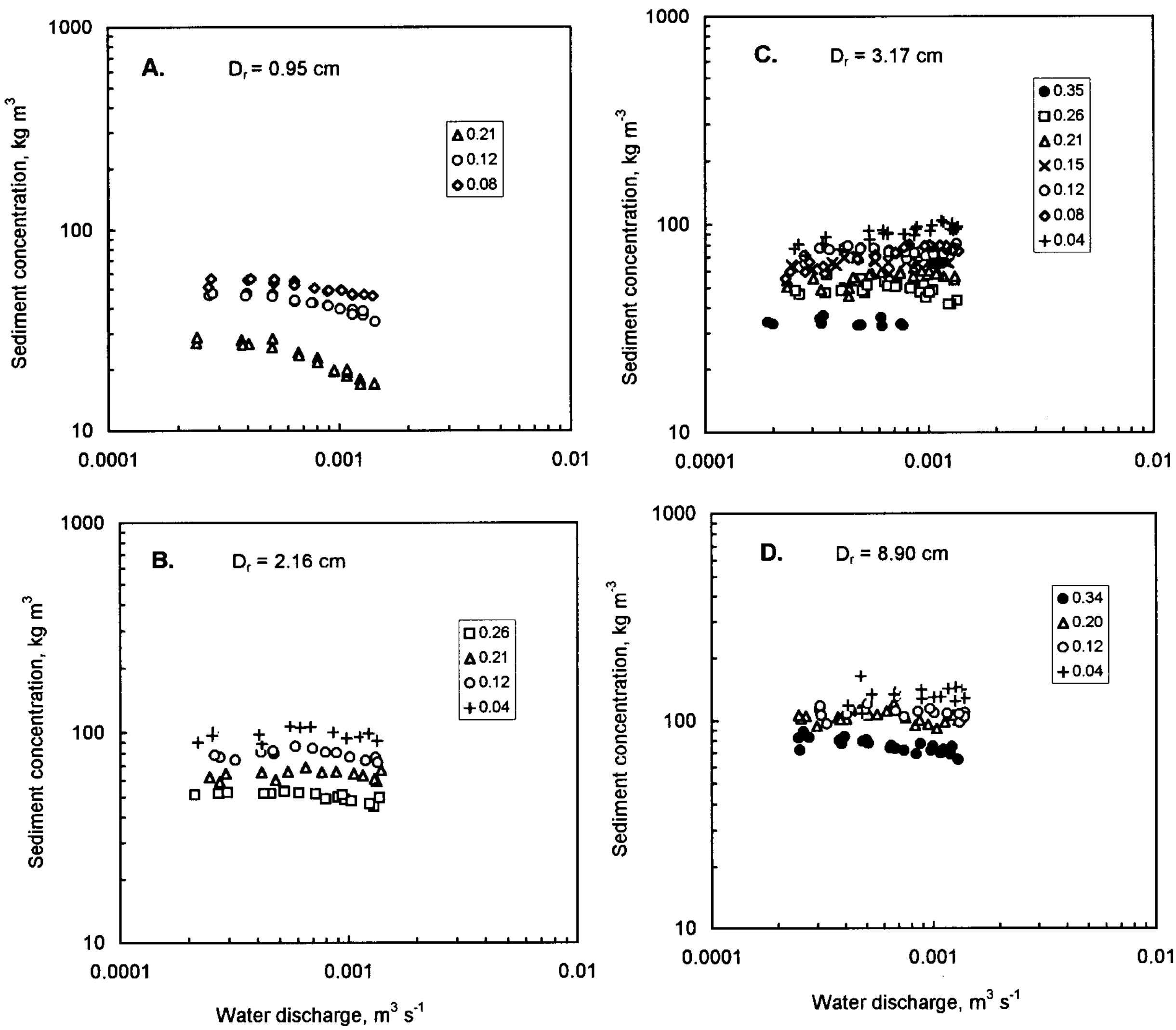

Figure 2. Graphs of sediment concentration against water discharge for cylinder-covered beds inclined at 5·5°. Median sediment size is 0·74 mm. Each graph displays data for a different cylinder diameter D_r. The different symbols signify different cylinder concentrations, as explained in the keys

variable was log T_w, and the initial independent variables were log $(\omega - \omega_c)$, log d, log u, C, D_r and I. Note that $(\omega - \omega_c)$, d and u are flow properties, C and D_r are roughness characteristics, and I is a rainfall property. The variables T_w, $\omega - \omega_c$, d and u were logged (to the base 10) because previous studies have established that the relations between transport capacity and $\omega - \omega_c$, d and u are power functions (e.g. Bagnold, 1977, 1980; Colby, 1964). The variables C, D_r and I were not logged because in some experiments they have a value of zero.

Before the regression could be performed the value of ω_c had to be determined for every experiment. In interrill flow the value of ω_c depends first and foremost on whether or not there is rainfall. Where rainfall accompanies interrill flow, raindrops continue to disturb the bed and lift particles into the flow as flow power diminishes and approaches zero. It follows that sediment transport takes place under rainfall no matter how small the flow power. Consequently, for the series 2 and 5 experiments which involved rainfall, ω_c was assumed to be zero. This assumption is supported by Moss *et al.*'s (1979) observations of rain-flow transportation on zero slopes.

Table II. Multiple regression analyses of immersed sediment transport capacity T_w on S, d and u, with S being included as a simple variable and as an element of the composite variables, τ, uS and ω. All variables were log-transformed in the analyses

	Regression coefficients						R^2	*SEE*
	S	u	d	τ	uS	ω		
3·883	1·612	1·905	0·535				0·903	0·120
–2·555		1·905	–1·078	1·613			0·903	0·120
3·883		0·293	0·535		1·612		0·903	0·120
–2·556		0·292	–1·079			1·613	0·903	0·120

For the remaining series without rainfall, ω_c was calculated in the following way. First, critical non-dimensional shear stress $\theta_c = \tau_c/[Dg(\rho_s-\rho)]$ was estimated from Miller *et al.*'s (1977) revision of Yalin's (1972) relation between θ_c and $\{[D^3g(\rho_s-\rho)]/\rho v^2\}^{1/2}$. Second, critical shear stress τ_c and critical flow depth d_c were calculated from

$$\tau_c = \theta_c Dg(\rho_s - \rho) \tag{5}$$

and

$$d_c = \frac{\theta_c D(\rho_s - \rho)}{\rho S} \tag{6}$$

Third, the critical Darcy–Weisbach friction factor f_c was computed. For the series 1 experiments this was done using the Keulegan (1938) equation

$$1/f_c^{1/2} = 2\cdot03\log(d_c/D) + 2\cdot21 \tag{7}$$

For the series 3, 4 and 6 experiments, multiple regression equations were derived for predicting f_c from C, D_r and S. Because f is largely determined by bed roughness, for a group of experiments with the same type of roughness and values of C, D_r and S, f remains unchanged as Q varies and is therefore equal to f_c for that set of bed roughness conditions. It follows that any equation that predicts f from C, D_r and/or S can be used to predict f_c. The derived equations are given in Table III. Fourth, critical flow velocity u_c was computed from

$$u_c = \left(\frac{8gSd_c}{f_c}\right)^{1/2} \tag{8}$$

Finally, ω_c was calculated using

Table III. Regression equations for estimating friction factor

Series	Regression equation	R^2
3	$\log f = 0{\cdot}045 + 2{\cdot}408\,C + 0{\cdot}343 \log S$	0·697
4	$\log f = -0{\cdot}296 + 1{\cdot}685\,C - 0{\cdot}966\,D_r$	0·473
6	$\log f = -0{\cdot}156 + 4{\cdot}171\,C$	0·511

$$\omega_c = \tau_c u_c \tag{9}$$

Although this method of computing ω_c is believed to give sound results, because $\omega_c << \omega$ in the great bulk of the experiments, any errors in ω_c will have little effect on the present regression analysis.

The regression analysis yielded the following equation:

$$\log T_w = -2{\cdot}638 + 1{\cdot}644 \log \omega - \omega_c) - 1{\cdot}037 \log d - 0{\cdot}577C + 2{\cdot}491D_r \tag{10}$$

with $R^2 = 0{\cdot}933$ and $SEE = 0{\cdot}100$ log units. All the regression coefficients are significantly different from 0 at the 0·01 level.

The first variable to enter the regression was $\omega - \omega_c$ which accounted for 66·8 per cent of the variance in T_w. Bearing in mind that $\omega \propto QS/w$, the strong positive correlation between T_w and ω is logically due to T_w increasing with Q and S. The second independent variable to enter the regression was d, which accounted for a further 23·0 per cent of the variation in T_w. The negative correlation between T_w and d reflects the fact that with $\omega - \omega_c$ being controlled by the regression, QS/w is more or less fixed, so any increase in surface roughness causes d to increase, u to decrease, and T_w to decrease. So d captures the effect of surface roughness on T_w. The third variable to enter the regression was C, and this variable added another 0·7 per cent to the explained variance. The negative correlation between T_w and C can be attributed to an increase in C causing an increase in d and a decrease in flow velocity and, hence, in T_w. The contribution to the explained variance is presumably small because this effect has already been largely captured by d. The final variable to enter the regression was D_r, which increased the explained variance by 2·8 per cent. The positive correlation between T_w and D_r indicates that, all other things (like C) being equal, small (i.e. narrow) roughness elements reduce sediment transport more than do large ones. There are two possible reasons for this. First, a large roughness element deflects a larger quantity of water than does a small one, thereby forming a larger horseshoe vortex which is able to sustain greater sediment transport (Bunte and Poesen, 1993). Second, large roughness elements, particularly in low concentrations, may actually concentrate the flow into well defined threads that are capable of transporting more sediment than a more dispersed flow through a comparable concentration of small elements.

Neither u nor I entered the regression. The absence of u is surprising given that in the analysis reported in Table II it contributed significantly (though modestly) to the explained variance in T_w. Its absence here indicates that it has no effect on T_w independently of $\omega - \omega_c$ and d. Evidently, $\omega - \omega_c$ and d are sufficient to account for all the variation in T_w due to flow properties. A number of studies have shown that rainfall increases sediment transport capacity (e.g. Walker *et al.*, 1978; Moss *et al.*, 1979; Guy *et al.*, 1987; Kinnell, 1991; Everaert, 1991). This increase is generally ascribed to raindrops detaching soil particles and lifting them into the flow (e.g. Kinnell, 1991) or to raindrops enhancing flow turbulence and thereby sustaining sediment transport (Guy *et al.*, 1987). These processes are most effective in

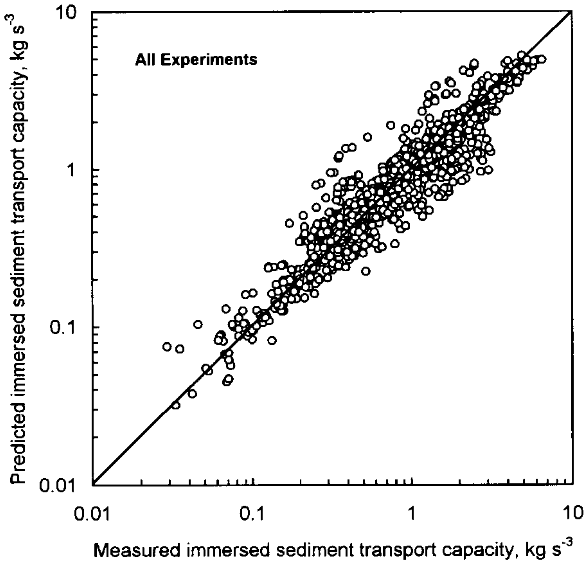

Figure 3. Graph of predicted against measured immersed sediment transport capacity for all 1506 flume experiments. The predictions are made using Equation 11

shallow laminar flow on smooth, gently inclined surfaces. The present experiments involve transitional and turbulent flow on rough surfaces with gradients up to 10°. The failure of I to appear in Equation 10 implies that under these conditions rainfall has no effect on transport capacity.

Probably the most significant feature of Equation 10 is that although the independent variables are able to explain 93·3 per cent of the variation in T_w, the proportion contributed by the surface roughness variables is only 3·5 per cent. The hydraulic variables $\omega - \omega_c$ and d account for fully 89·8 per cent of the total variance. If the surface roughness variables are discarded, regression analysis yields the equation

$$\log T_w = -3 \cdot 038 + 1 \cdot 726 \log(\omega - \omega_c) - 1 \cdot 212 \log d \tag{11}$$

with $R^2 = 0{\cdot}898$ and $SEE = 0{\cdot}123$ log units. In Figure 3 T_w predicted by Equation 11 is plotted against measured T_w for all 1506 experiments. The points scatter symmetrically around the line of perfect agreement, suggesting that Equation 11 gives unbiased predictions of T_w. In Figure 4, graphs of predicted T_w against measured T_w for the individual series afford a more stringent test of Equation 11. These graphs reveal minor biases, notably in series 6. Even so, considering the wide range of experimental conditions, this is a remarkable result which highlights the theme of this study, namely that regardless of the size, shape and concentration of the roughness elements, their effect on sediment transport capacity is largely captured by the hydraulic variables $\omega - \omega_c$ and d.

Finally, as noted above, EUROSEM predicts the transport capacity of interrill flow on rough surfaces using an equation based on Everaert's (1991) experiments on plane beds. This equation contains Govers' (1990) effective flow power $(\omega - \omega_c)^{1 \cdot 5}/d^{0 \cdot 667}$ raised to a power that depends on sediment size. In contrast to Everaert's experiments, the present experiments were conducted on rough as well as plane beds. Equation 11, which summarizes these experiments, indicates that the ratio of the $\omega - \omega_c$ exponent to the d exponent is not always $1{\cdot}5/{-0{\cdot}667} = -2{\cdot}250$, as it is in the EUROSEM equation, but at least for 0·74mm sand it is $1{\cdot}726/{-1{\cdot}212} = -1{\cdot}424$. Additional experiments are needed for a variety of sediment sizes to gain a fuller understanding of the behaviour and controls of the $\omega - \omega_c$ and d exponents.

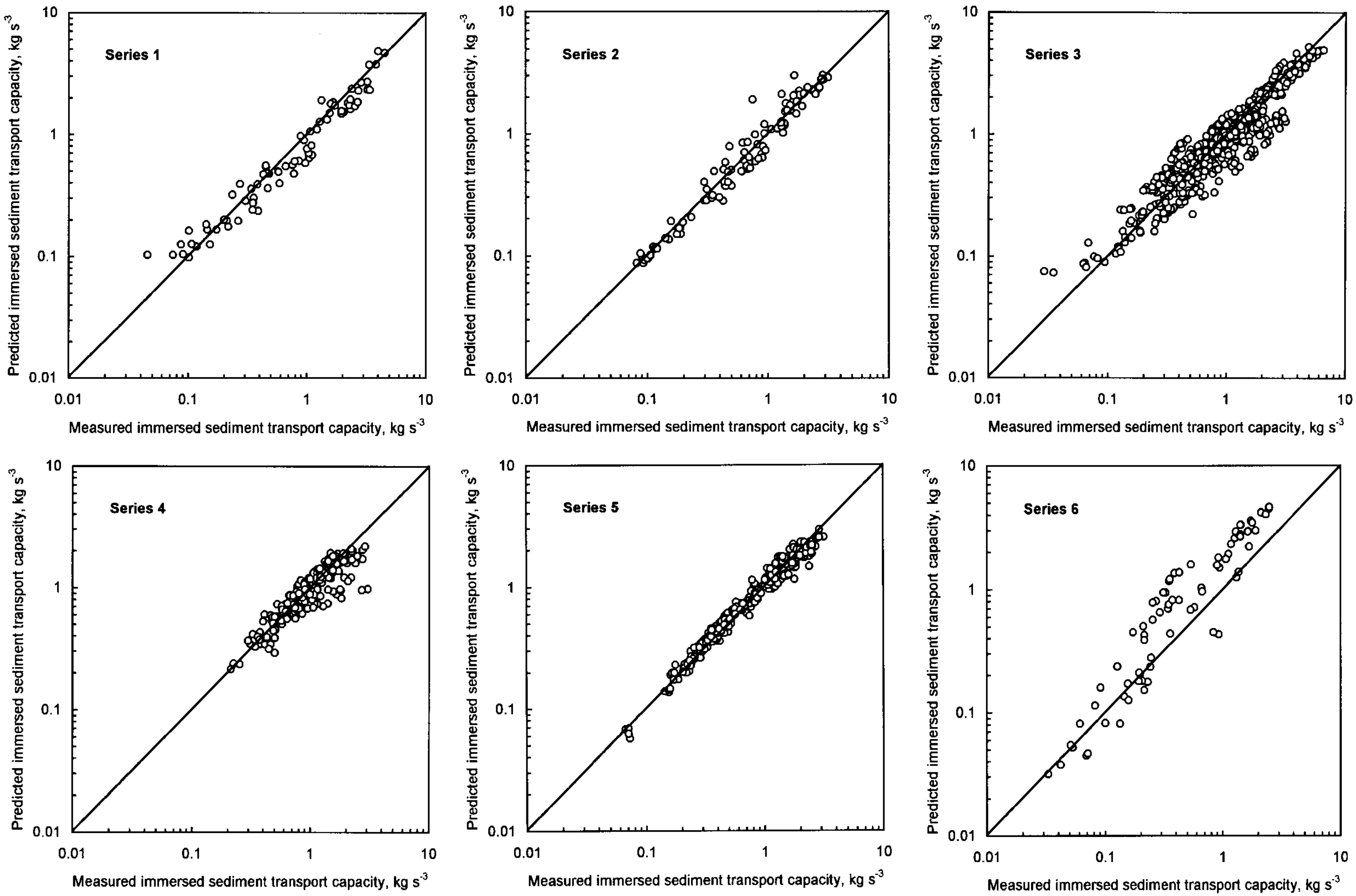

Figure 4. Graphs of predicted against measured immersed sediment transport capacity for the six individual series of experiments. The predictions are made using Equation 11

CONCLUSION

In this study it has been shown that there is no need to resort to shear stress partitioning to predict sediment transport capacity of interrill overland flow on rough surfaces. Good predictions can be obtained for transitional and turbulent flows using the hydraulic variables excess flow power and flow depth. Evidently, flow depth, when combined with excess flow power, largely captures the effect of surface roughness on transport capacity. This finding promises to simplify greatly the task of developing a general sediment transport equation for interrill overland flow on rough surfaces. Should such an equation be developed, it will require accurate information on flow hydraulics. Thus, the problem of predicting transport capacity of interrill flow demands not only a sediment transport equation but a reliable means of measuring or modelling interrill flow hydraulics on rough surfaces.

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